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Abstract

Recent measurements of the volatile composition of the coma of Comet 67P/Churyumov-Gerasimenko (hereafter 67P) allow constraints to be set on the origin of volatile elements (water, carbon, nitrogen, noble gases) in inner planets' atmospheres. Analyses by the ROSINA mass spectrometry system onboard the Rosetta spacecraft indicate that 67P ice has a D/H ratio three times that of the ocean value (Altwegg et al., 2015) and contains significant amounts of N_2, CO, CO_2, and importantly, argon (Balsiger et al., 2015). Here we establish a model of cometary composition based on literature data and the ROSINA measurements. From mass balance calculations, and provided that 67P is representative of the cometary ice reservoir, we conclude that the contribution of cometary volatiles to the Earth's inventory was minor for water (≤ 1%), carbon (≤ 1%), and nitrogen species (a few % at most). However, cometary contributions to the terrestrial atmosphere may have been significant for the noble gases. They could have taken place towards the end of the main building stages of the Earth, after the Moon-forming impact and during either a late veneer episode or, more probably, the Terrestrial Late Heavy Bombardment around 4.0-3.8 billion years (Ga) ago. Contributions from the outer solar system via cometary bodies could account for the dichotomy of the noble gas isotope compositions, in particular xenon, between the mantle and the atmosphere. A mass balance based on ^{36}Ar and organics suggests that the amount of prebiotic material delivered by comets could have been quite considerable – equivalent to the present-day mass of the biosphere. On Mars, several of the isotopic signatures of surface volatiles (notably the high D/H ratios) are clearly indicative of atmospheric escape processes. Nevertheless, we suggest that cometary contributions after the major atmospheric escape events, e.g., during a Martian Late Heavy Bombardment towards the end of the Noachian era, could account for the Martian elemental C/N/^{36}Ar ratios, solar-like krypton isotope composition and high ^{15}N/^{14}N ratios. Taken together, these observations are consistent with the volatiles of Earth and Mars being trapped initially from the nebular gas and local accreting material, then progressively added to by contributions from wet bodies from increasing heliocentric distances. Overall, no unified scenario can account for all of the characteristics of the

inner planet atmospheres. Advances in this domain will require precise analysis of the elemental and isotopic compositions of comets and therefore await a cometary sample return mission.

1. Introduction

According to cosmochemical constraints and simulations, the inner planets grew first "dry", and water and atmophile elements were contributed later on by volatile-rich objects originating from larger heliocentric distances (Morbidelli et al., 2000; Raymond et al., 2004; Albarède et al., 2013; Marty, 2012). A cometary origin for the Terrestrial oceans, the atmosphere and the organics is often advocated (Oro, 1961; Anders and Owen, 1977; Safronov, 1972) because (i) comets are volatile-rich as they contain abundant organics and ices (30-50 wt. % H_2O), and (ii) their high eccentricities and hyperbolic trajectories increase their probability to encounter inner solar system bodies. Other potential sources of water are "wet" asteroids. Meteorites are generally dry, although some of them, the carbonaceous chondrite clan, can contain up to 15 wt. % equivalent H_2O as hydrated phases. Because these phases are remnant of water-rock interactions on asteroids, the original water content could have been significantly higher, possibly matching those of comets. This would argue in favor of a continuum between meteoritic and cometary compositions in the past. The Terrestrial oceans (1.5×10^{24} g) correspond to 0.025 % of the Earth's mass. Taking into account water in the deep Earth may raise the Terrestrial H_2O inventory to a maximum of 0.2 wt. % (Marty, 2012). Assuming conservation of volatile elements during impacts, that is, no loss of volatiles to space (an assumption discussed later on), Terrestrial water could have been supplied by a few wt. per mil to a few wt. percent of cometary, or asteroidal, material to a dry proto-Earth, leaving the bulk chemical composition of Earth little affected.

Dynamical models suggest that, during its formation, the Earth accreted water and other volatile elements predominantly from planetesimals parented to primitive asteroids, i.e. with composition similar to carbonaceous chondrites (Morbidelli et al., 2000; Raymond et al. 2004, 2007; O'Brien et al., 2006, 2014). Comets, intended as planetesimals of trans-Neptunian origin, should have had a negligible contribution during Earth's accretion. Instead, once the Earth's formation was completed,

77 possibly several hundreds of million years (Ma) afterwards, the Earth should have accreted non-
78 negligible amounts of mass from asteroids and/or from icy bodies of trans-Neptunian orgin, during a
79 dynamical instability event that re-structured the orbital configuration of the giant planets (the so-
80 called "Nice model"; Gomes et al., 2005). These models predict contributions of icy bodies in
81 proportions comparable to those of asteroids. From the lunar impact crater record, Chyba (1990)
82 argued that the Earth could have acquired an exogenous ocean around 4 Ga ago if comets comprised
83 ≥ 10 wt. % of the impacting population. Marty and Meibom (2007) made the point that, if comets
84 contain noble gases in amount predicted by laboratory ice trapping experiments (see below), then the
85 fraction of cometary material in the late heavy bombardment on Earth could not have been more than
86 0.5 %, as constrained by the noble gas inventory of the atmosphere.

87 Cometary contributions might have left an imprint in the composition of the ocean-atmosphere.
88 Owen and Bar-Nun (1995) noted that the noble gas elemental composition of the atmospheres of
89 Mars and Earth (which are different from those of chondrites by having a one order of magnitude
90 lower Xe/Kr ratio, Fig. 1), are consistent with a cometary origin. Experiments of noble gas trapping
91 in growing amorphous ice have shown that the absolute amounts and their relative proportions
92 depend on the trapping kinetics, pressure and, importantly, temperature (Bar-Nun et al., 1985). Owen
93 and Bar-Nun (1995) proposed a cometary ice formation temperature of 50 K to account for the
94 Martian and Terrestrial Ar/Kr/Xe ratios. Other works questioned the relevance of these experimental
95 conditions to those of the protosolar nebula (Notesco and Bar-Nun, 2005; Yokochi et al., 2012) or
96 preferred trapping of cometary noble gases in clathrates rather than in amorphous ice (Mousis et al.,
97 2010). Importantly, because Xe from the protosolar nebula is expected to be quantitatively trapped in
98 cometary ice (Notesco and Bar-Nun, 2005; Yokochi et al., 2012; Mousis et al., 2010) and because
99 physical adsorption does not induce significant Xe isotope fractionation (Marrocchi and Marty,
100 2013), cometary Xe should have a solar-like isotopic composition. However, the xenon isotopic
101 compositions of the atmospheres of Earth and Mars are not solar (nor chondritic), being enriched in
102 the heavy isotopes relative to the light ones by 3-4 % par atomic mass unit (Fig. 1). Taking into

account this discrepancy, Dauphas (2003) proposed a dual origin for the Terrestrial atmosphere. In his model, a primary atmosphere of solar composition was lost to space, leaving residual gases mostly made of isotopically fractionated Xe. Then comet-like bodies contributed noble gases depleted in Xe (following the same argument as Owen and Bar-Nun, 1995, that cometary ice should depleted in Xe compared to Chondritic and other noble gases, Fig. 1). This model accounts successfully for the elemental and isotopic compositions of atmospheric noble gases and for the "xenon paradox": Atmospheric xenon is depleted relative to Chondritic Xe but also relative to atmospheric Kr, and enriched in its heavy isotopes relative to any known cosmochemical composition (Fig. 1), both properties being paradoxical if xenon depletion was due atmospheric escape. However, the Xe paradox might be related to a different process and may not be a diagnostic of source composition. Indeed, from the analysis of ancient (>2.7 Ga-old) minerals, Pujol et al. (2011) argued for continuous loss and isotopic fractionation of atmospheric xenon to space through geological time as a cause of the peculiar composition of atmospheric Xe. Therefore the xenon paradox might be due to secondary long-term processing and not to a source effect.

The hydrogen and nitrogen isotopes give independent insight into the origin(s) of planetary volatiles (e.g. Alexander et al., 2012; Marty, 2012; Füri and Marty, 2015; Fig. 2). Taken together, they suggest an asteroidal, rather than cometary, origin for nitrogen and water in the Terrestrial atmosphere and the oceans. The D/H ratios of comets are generally higher by a factor of 2-3 relative to the ocean's value and to most of the chondritic values, prohibiting a genetic relationship between cometary ice and Terrestrial water. Available data are mostly from long-period comets thought to originate from the Oort cloud. The measurement of an ocean-like D/H ratio for a Jupiter family comet (Hartogh et al., 2011), presumably originating from the Kuiper belt, suggests a D/H heterogeneity among comet families. Recently, a D/H value three times the ocean's has been measured in the coma from Comet 67P/Churyumov-Gerasimenko (thereafter 67P), also a Jupiter family member (Altwegg et al., 2015). The nitrogen isotopes give a similar story. Chondrites have variable N isotope ratios which distribution coincides with the Terrestrial composition. In contrast,

129 comets have $^{15}N/^{14}N$ ratios about two times the Terrestrial and bulk chondritic values (Fig. 2). One
130 possible caveat is that only HCN, CN, and NH_2^+ could be measured at distance by spectroscopy
131 (there is no in-situ measurement on a comet), and it is not clear if these nitrogen species are
132 representative of bulk cometary N.

133 In this contribution we use the recent compositional data of the 67P coma obtained by the
134 ROSINA mass spectrometer on board of the ROSETTA spacecraft to set constraints on the origin(s)
135 of volatile elements on Earth and Mars. Importantly, this instrument measured for the first time a
136 noble gas - argon - in cometary gases (Balsiger et al., 2015). We present first a model composition
137 for cometary matter using literature data and assuming that 67P is representative of the cometary
138 reservoir. We then attempt a mass balance to infer the possible contribution of the outer solar system
139 to the Terrestrial and Martian inventory of water, carbon, nitrogen and noble gases.

140 **2. Reservoir inventories**

141 *2.1. Comets*

142 *2.1.1. Elemental abundances*

143 The chemical and isotopic compositions of comets is constrained by (i) in-situ measurements by
144 spacecrafts (e.g., A'Hearn et al., 2011; Le Roy et al., 2015; Jessberger et al., 1988; Kissel and
145 Krueger, 1987) (ii) spectroscopic observations of comae (e.g., Bockelée-Morvan, 2011; Mumma and
146 Charnley, 2011), and (iii) laboratory analysis of cometary grains recovered by the Stardust mission
147 (Brownlee, 2006) and of interplanetary dust particles presumably of cometary in origin (e.g., Duprat
148 et al., 2010).

149 Comets are made of refractory silicates and metal, refractory and non-refractory organics, and ices
150 (Jessberger et al., 1988; ; Kissel and Krueger, 1987; Mumma and Charnley, 2011; Bockelée-Morvan,
151 2011). The analysis of cometary grains recovered by the Stardust mission revealed the occurrence of

several high temperature phases reminiscent of chondritic material. These are calcium-aluminium rich inclusions, chondrules and refractory olivines. Stable isotope compositions of these phases indicate an inner solar system origin for this material (Brownlee, 2006). From mass balance, silicate + metal (Si, Mg, Fe) grains make about 25-30 wt. % (Jessberger et al., 1988). Refractory organics constitute a significant fraction of C and, to a lesser extent, of N and H, and must be taken into account in the bulk composition of cometary matter (Greenberg, 1998). This is confirmed by a refractory organics/silicate+metal ratio close to 1 for dust emitted by comet Halley (Kissel and Krueger, 1987). Therefore, organic refractory particles (the so-called CHON grains, Jessberger et al., 1988) could make another 25-30 wt. %, and non-refractory carbon-bearing species could contribute about 4-8 wt. % (Greenberg, 1998). The proportion of ice amounts for 30-50 wt. %. The composition of comae (Table 1) gives insight into the ice composition (~ 80 wt. % H_2O with CO, CO_2, CH_3OH, CH_4, H_2S, and N-compounds including CN, HCN, NH_3 and N; Bockelée-Morvan, 2011; Hässig et al., 2015; Rubin et al., 2015).

Noble gases are able to set stringent constraints on the origin of planetary volatiles on one hand, and on processes and conditions of comet formation on another hand. Unfortunately the abundances of noble gases in comets are essentially unknown as these elements are extremely difficult to detect by spectroscopy. Stern et al. (2000) proposed an Ar/O ratio of $1.8\pm1.0 \times 10^{-4}$ for Comet Hale Bopp, lower than the solar value of 46×10^{-4} (Weaver et al., 2002, and refs. therein), but this estimate has not been confirmed. Weaver et al. (2002) suggested that the cometaryAr/O ratios are lower than 42×10^{-4}, based on three comet observations. The only reliable set of measurements are those done on 67P coma using the ROSINA mass spectrometer which directly the 67P coma composition (Balsiger et al., 2015). They indicate a $^{36}Ar/H_2O$ varying between 0.1×10^{-5} and 2.3×10^{-5} during the 3 days measurement period. Considering that Ar/O ~1.3 Ar/H_2O (Weaver et al., 2002), and taking into account a solar isotope composition for argon ($^{40}Ar \sim 0$, $^{38}Ar/^{36}Ar = 0.183$), the corresponding Ar/O range of 67P is $0.01\text{-}0.35 \times 10^{-4}$, thus lower than the upper limits given above, and 2-3 orders of magnitude lower than the solar ratio of 46×10^{-4}. It is worth mentioning that even if noble gases are

178 mostly hosted in ices, refractory grains may also contain large amounts of these elements, as
179 suggested by the discovery of high He and Ne abundances in grains of comet Wild2 recovered by the
180 Stardust spacecraft (Marty et al., 2008). However, a mass balance is not possible due to the lack of
181 relevant Ar data, and we adopt here the 67P coma composition.

182 *2.1.2. Isotopes*

183 The stable isotope ratios relevant to this study are the D/H ratios of comets, with special reference
184 to the 67P value ([5.3 ± 0.7] x 10^{-4}, ~ 3 times the ocean water ratio; Altwegg et al., 2015), and N
185 isotope variations depicted in Fig. 2, althougth no data are yet available for 67P. For noble gases,
186 only the $^{36}Ar/^{38}Ar$ ratio has been measured in the coma of 67P (5.4 ± 1.6; Balsiger et al., 2015),
187 consistent with a solar system composition, but with a precision too low to permit identification of a
188 specific cosmochemical component. Models of ice formation from a protosolar nebula gas predict
189 that the isotopic composition of noble gases are solar. However, there is considerable uncertainty
190 arises for the mode of trapping (i.e., neutral versus ionized) which may fractionate, or not, noble gas
191 isotopes, and for the local composition of the gas during ice formation. Furthermore, the analysis of
192 Stardust refractory grains revealed a high abundance of light noble gases and has shown that the
193 neon isotope composition is closer to a meteoritic signature than to a solar one, precluding to
194 establish a definite composition for cometary noble gases. Measurements of noble gas isotopes in
195 comets should be a given a high priority in future space programmes. In the absence of such, we only
196 discuss here the case of argon. The model cometary composition is given in Table 2. The
197 uncertainties represent the range of values observed in comets, unless a single set of values is
198 available.

199 *2.2 Carbonaceous chondrites*

200 For noble gases, we use a literature survey of 87 noble gas analyses of CI and CM (Schultz and
201 Franke, 2004). The H, C and N data are from 40 analyses compiled by Kerridge (1985). The

respective mean chondritic abundances of C, N, H_2O and ^{36}Ar are, in mol/g: 2.3±1.2 x 10^{-3}, 6.4±2.5 x 10^{-5}, 3.7±1.0 x 10^{-3} and 3.1±1.6 x 10^{-11}, respectively.

2.3 Earth

The inventory of volatile elements at the Earth's surface (the atmosphere, the oceans and the crust) is reasonably well known. The problem is to estimate the amount of volatiles in the mantle, since magmas tend to degas readily upon eruption, and the inferred source compositions are model-dependent. Recent estimates of the bulk Earth (mantle plus surface) volatile inventories have been given by Marty (2012), Halliday (2013) and Dauphas and Morbidelli (2014). We adopt here two limiting cases. The lower limit is based on the reconstruction of the initial volatile inventory of the mantle regions sampled by volcanism. The initial contents of water, C, N and noble gases in the mantle source of mid-ocean ridge basalts (MORB; volumetrically the main source of lavas at the Earth's surface) is derived from a combination of lava fluxes and calibration to relevant tracers like ^3He or refractory elements like Ce, Nb. Halliday (2013) considered also the source of Ocean Island basalts (another important source of volcanism thought to tap deep regions of the mantle), ("HA" in Figs. 3 and 4). Dauphas and Morbidelli (2014; "DM") obtained an inventory comparable to the one of Halliday (2013) from global calibration of volatiles to refractory (Ce, Nb) elements, and obtained a total bulk Earth water content of about 3 ocean masses (M_{Ocean} equivalent to 230 ppm H_2O for bulk Earth). The isotopic compositions of helium, neon, argon and xenon suggest that deep mantle regions could be richer in volatiles than the convective mantle sampled by mid-ocean ridges. K-Ar systematics indicate that about half of radiogenic ^{40}Ar produced in Earth over 4.5 Ga is still trapped in the solid Earth, the rest having been degassed into the atmosphere (Allègre et al., 1996). Calibrating volatile elements to ^{40}Ar trapped in the mantle led Marty (2012) to propose a bulk mantle ("MA") higher than the above estimates, corresponding to 4-10 ocean masses. This approach does not rely on models of the structure and the composition of mantle domains. However, it assumes a chondritic composition for Terrestrial refractory elements. This is because the potassium content of

227 the Earth (Arevalo et al., 2009) is calibrated against refractory uranium as potassium is semi-volatile
228 and its abundance cannot be retrieved from meteoritic analogs. A chondritic Earth composition for
229 refractory elements has been recently questioned on the basis of sub-chondritic Terrestrial Nd
230 isotope ratios, with the possibility that up to half of U, and therefore K, could have been stripped off
231 by impacts on a differentiated Earth (e.g., Jellinek and Jackson, 2015; and refs. therein). This would
232 reduce the ^{40}Ar inventory and therefore allow much lower volatile contents. For this reason, we
233 consider that the MA inventory of Table 4 gives an upper limit for the Terrestrial volatile budget.

234 **3. Discussion**

235 *3.1. How representative are cometary data?*

236 We assume here that 67P data on one hand, and available cometary data on another hand, are
237 representative of the composition of icy bodies that scattered in the inner solar system. This
238 assumption is certainly fragile in the sense that key species measured in 67P coma are likely to vary
239 from comet to comet. However, comets formed probably from comparable building material, water
240 and other ice species making ~30-50 % of total mass. For most volatile species given in Tables 2 and
241 3, measurements were done spectroscopically on several comets (Mumma and Charnley, 2011,
242 Bockelée-Morvan, 2011, Le Roy et al., 2015) so the large range of observed values is considered to
243 be representative of the natural variability.

244 For noble gases, the Ar measurement of Balsiger et al. (2015) is the only available set of data, but
245 we propose below that the range of observed values is representative of the coma composition.
246 Fractionation of a minor, low temperature species/element like argon relative to H_2O can range over
247 1-2 order(s) of magnitude, depending on its sublimation temperature relative to that of H_2O ice.
248 Marboeuf and Schmitt (2014) modeled fractional degassing of cometary volatiles as a function of the
249 distance to the perihelion, under a range of different assumptions concerning the thermal inertia of a
250 comet, the nature of ice, and the abundance and distribution of the different volatile species. Argon

was not considered in their study, so we take as a proxy CO which has a comparable condensation/sublimation temperature. Results of their modeling indicate that, at a distance between 3 and 4 AU from the perihelion, the CO/H_2O ratio of the coma is comparable to that of the bulk comet. Argon measurements by the ROSINA mass spectrometer were done on October 19, 20, 22, and 23, 2014 when Comet 67P was at ~3.1 AU from the Sun, thus possibly minimizing fractionation of minor species like Ar relative to H_2O. The 67P $^{36}Ar/H_2O$ ratio of 0.1-2.3 x 10^{-5} is also comparable to, although lower than, ratios predicted from modeling of ice trapping experiment data (2-40 x 10^{-5} in the 30-45 K range according to Dauphas, 2003; 8-75 x 10^{-5} according to Marty and Meibom, 2007, based on experimental data from Notesco and Bar-Nun, 2005). Models and experiments of noble gas trapping into cometary ice indicate that the main controlling factors on the abundance of trapped noble gases are the temperature and the partial pressures of gases (Bar-Nun et al., 1985; Yokochi et al., 2012), so that other parameters of comet formation being equal, we assume that the Ar content of cometary matter is represented by 67P data. Although the above assumptions may constitute the main weak point of the present approach, we note that uncertainty range covers one order of magnitude of the stated values (the $^{36}Ar/H_2O$ ratio) whereas the present conclusions here are based on variations of key parameters over several orders of magnitude.

3.2. Impact driven atmospheric loss

The present approach assumes that impacts are conservative, that is, contributed volatiles and target volatiles are not lost into space. This was likely not the case since giant impacts probably led to significant atmospheric loss (e.g., Genda and Abe, 2005). Recent models argue for a steady state regime between replenishment and atmospheric escape, depending on velocities, atmospheric entry angles, and sizes of impacting bodies (De Niem et al., 2012; Schlichting et al., 2014), but modeling the overall budget of volatile loss/gain is presently out of reach. Instead, we turn to a mass balance approach involving radiogenic ^{129}Xe, produced from the extinct radioactivity of iodine-129 ($T_{1/2}$ = 16 Ma). Iodine was present when the Earth formed, as shown by the occurrence on Earth of the stable

276 isotope ^{127}I. From meteoritic studies, it is possible to derive the solar system initial $^{129}I/^{127}I$ (1.0×10^{-4}). If the Earth formed when ^{129}I was still alive, then terrestrial reservoirs should present excesses of ^{129}Xe produced by the decay of ^{129}I. This is the case as an excess of 7% ^{129}Xe (the main reservoir of terrestrial xenon) is present in air. This excess corresponds to about 1-2 % of ^{129}Xe that could have been present if the Earth formed at the start of solar system formation. Assuming that produced ^{129}Xe was lost before the atmosphere became closed for volatile escape, one obtains a I-Xe closure age of ~100 Ma after start of Solar System formation for the terrestrial system (Wetherhill 1980). Atmospheric xenon including ^{129}Xe might have been escaping from the terrestrial atmosphere for prolongated periods of time, possibly during about 2 Ga (Pujol et al., 2011), so that the ^{129}Xe excess has to be corrected for subsequent loss of atmospheric Xe. After correction, the closure age of the atmosphere becomes 30-60 Ma (Avice and Marty, 2014). A comparable budget arises when combining extinct radioactivities of ^{129}I and ^{244}Pu ($T_{1/2}$ = 82 Ma), implying that Xe loss was early, within a few tens of Ma after start of Solar System formation. Thus, according to I-Pu-Xe mass balance, early atmospheric loss, presumably during the proto-Earth building epoch, might not have exceeded 60-80 % of atmospheric Xe, and probably less after the Moon-forming impact which occurred within 40-90 Ma after start of Solar System formation. We tentatively conclude that the atmosphere was grossly conservative, within a factor of two, after the Moon-forming impact (the period of time relevant to cometary contributions, see next sub-sections).

3.3. Origin of major volatile species on Earth

The Ar content of cometary material inferred from the analysis of argon in 67P coma is 3-4 orders of magnitude higher than the one of chondritic material, which makes Ar, and by inference other heavy noble gases, a key tracer of cometary contribution to Earth. Together with the D/H ratio, the ratio between argon and H_2O, C and N permits to establish firm constraints on the contribution of 67P-like material to Terrestrial volatiles. Swindle and Kring (2001) considered the ratio between ^{36}Ar and H_2O in comets (they took the Hale Bopp's $^{36}Ar/O$ ratio from Stern et al., 2000, which has been

301 questioned since then) to infer an extremely low contribution of cometary material to the surfaces of
302 Earth and Mars. Dauphas (2003) noted that the low Xe/H_2O ratio of the Earth compared to other
303 cosmochemical end-members could set stringent constraints on the maximum amount of cometary
304 water on Earth. Since then, it has been proposed that the low terrestrial Xe/H_2O ratio could be
305 secondary (Pujol et al.. 2011), making the argument less constraining. Figs. 3 and 4 represent mixing
306 curves between chondritic and cometary end-members using the $^{36}Ar/X$ (where X represents water,
307 nitrogen or carbon) ratio versus the D/H ratio. In fact the ^{36}Ar concentration is more sensitive than
308 the D/H ratio alone, because contribution of 67P-like material can strongly impact the atmospheric
309 inventory of argon. For water, a cometary contribution of the order of 1% or less is allowed
310 depending on the bulk silicate Earth + surface composition (Fig. 3a), confirming that Terrestrial
311 water is mostly non-cometary. A comparable conclusion arises for nitrogen (Fig. 3b), although a
312 larger cometary contribution up to a few percent is allowed, due to the depletion of N in Earth
313 (Marty, 2012). Nitrogen isotopes may permit to get independent constraints because spectroscopic
314 measurements on comets indicate that the measured N-bearing species, CN, HCN and NH_2 (the latter
315 presumably from NH_3) are all enriched by a factor of 2 in ^{15}N relative to ^{14}N (e.g., Füri and Marty,
316 2015, for a review). However, other N species like N_2 (Rubin et al., 2015) and refractory organics are
317 present in comets with unknown N isotope composition, preventing to establish an isotope budget for
318 the cometary reservoir.

319 Based on the same approach, Terrestrial carbon appears to originate from chondrite-like material
320 (Fig. 4a), with a possible cometary contribution being lower than ~1 %. This conclusion stands even
321 when considering that cometary carbon may be hosted not only in ice but also in refractory organics
322 from the nucleus.

323 *3.4. Possible cometary origin for atmospheric noble gases*

324 Contrary to major volatiles, the mixing diagram of Fig. 4b suggests that atmospheric argon, and
325 by extension other heavy noble gases, could originate from a cometary reservoir, a possibility

already proposed by Anders and Owen (1977) and Owen and Bar-Nun (1995) based on the noble gas abundance pattern of the atmosphere, and by Marty and Meibom (2007) from mass balance considerations. For the bulk silicate Earth inventory, the conclusion depends on the adopted composition. For the HA and DM inventories, a major cometary origin for argon is clearly possible (Fig. 4b). The MA inventory does not require contribution of cometary noble gases to the bulk Earth (Fig. 4b). In this model, noble gases are mostly in the atmosphere, and major volatiles (H_2O, C) are in the mantle, the sum of which being consistent with the addition of ~ 2 wt. % of carbonaceous chondrite type material.

3.5. Potential cometary contribution during the major bombardment periods

Here we examine possibilities of a cometary contribution from the dynamical point of view. From solar system dynamics and isotope cosmochemistry, the major events that contributed potentially wet material to the building of Earth are:

1- Contribution of wet bodies to the growing proto-Earth (Morbidelli et al., 2000; Raymond et al., 2004; O'Brien et al., 2006, 2014). Dry material in the Terrestrial region became scarce as a result of accretion to larger bodies, leaving space to wet contributions from larger heliocentric distances. These late contributions should have consisted of volatile-rich asteroids, rather than comets. In line with this, krypton (Holland et al., 2009), hydrogen and nitrogen isotopes (Alexander et al., 2012; Marty, 2012) point to a chondritic, rather than cometary, source for contributed volatiles that are now stored in the Terrestrial mantle (Fig. 2).

2- The giant impact that led to the Moon formation. It is not clear if this event resulted or not in a major loss of volatile elements the proto-Earth. Already existing atmosphere and oceans could have survived the event (Genda and Abe, 2005), and/or the impactor could have contributed volatiles to the Earth-Moon system. From key isotope signatures (e.g., oxygen), the impactor was likely to originate from the iner solar system, and a cometary contribution is not considered further.

350 3- A late veneer (LV) might have supplied chondritic material after the last equilibration of
351 mantle material with the core, presumably after the Moon forming impact (Kimura et al., 1974). Its
352 contribution is estimated to be 0.3 wt. % (range 0.1-0.8 %) of the Terrestrial mass (Righter, 2003),
353 that is, $\sim 2 \times 10^{25}$ g. According to dynamical models, the LV was made of planetesimal leftover from
354 the Terrestrial planet formation, without contribution of comets (Bottke et al., 2010; Raymond et al.
355 2013; Morbidelli et al., 2012; Jacobson et al., 2014). In line with this view, a 100% cometary LV
356 (Fig. 5a) contributing 0.3 % of the Terrestrial mass would supply too much atmospheric ^{36}Ar by 2-3
357 orders of magnitude, and would have dominated the H, C and N inventory, which is unlikely. A
358 cometary contribution of 1 % or less could account for atmospheric Ar without impacting the stable
359 isotope signatures (Fig 5b).

360 4- The Terrestrial late heavy bombardment (TLHB) is an extrapolation of the late heavy
361 bombardment thought to have occurred on the Moon from the size-age distribution of lunar craters
362 (Tera et al., 1974). Estimates of the mass contribution to Earth during the TLHB, based on the Lunar
363 cratering record, range from 2.2×10^{23} g (Hartmann et al., 2002) down to 4×10^{22} g (assuming that
364 the TLHB started at the time of formation of Nectaris basin; Morbidelli et al., 2012). From a
365 modeling approach, Levison et al. (2001) proposed a mass contribution of 8×10^{22} g. Gomes et al.
366 (2005) estimated a TLHB of 4×10^{23} g, about half of which being cometary. The newest version of
367 the Nice model, however, reduces the mass flux to the Earth to about 4×10^{22} g of asteroidal material
368 (Bottke et al., 2012) and 2-4 $\times 10^{22}$ g of cometary material (depending on the dynamical state of the
369 trans-Neptunian disk before the dynamical instability). For the calculations presented here, a
370 conservative value of 2.0×10^{23} g is adopted. Marty and Meibom (2007) concluded that less than 1 %
371 comets were present in the TLHB, the rest being asteroidal material, and proposed that atmospheric
372 noble gases could constitute the only remaining evidence for the TLHB. Their mass balance was
373 based on noble gas cometary concentration from ice trapping experiments, which are higher than the
374 one constrained by Ar in 67P. With the ROSINA ^{36}Ar measurement, a 100% cometary TLHB would
375 supply too much ^{36}Ar to the atmosphere (Fig. 5c). It would also supply most major volatiles at the

376 Earth's surface, which is not allowed by the D/H and $^{15}N/^{14}N$ ratios of the atmosphere and oceans. A
377 cometary fraction of ~10% in the TLHB accounts reasonably well for atmospheric ^{36}Ar without
378 impacting the stable isotope budget (Fig. 5d). Thus a cometary contribution of ~2.0 x 10^{22} g respects
379 our mass balance and is in excellent agreement with the latest model results.

380 5- The post TLHB continuous flux of interplanetary dust particles (IDPs), and micrometeorites is
381 estimated at ~2 x 10^{20} g, assuming a constant flux comparable to modern one (about 30,000 tons/yr,
382 Love and Brownlee, 1993). A near-constant flux (with a possible factor of 5 increase in the last 0.5
383 Ga) is consistent with the lunar soil record of extraterrestrial material (Culler et al., 2000; Hashizume
384 et al., 2002). From the analysis of IDPs, this flux consists of both cometary and asteroidal materials.
385 A 100% continuous IDP flux would not impact the volatile inventory of the Earth's surface including
386 noble gases, even when this flux is increased by a factor of 5 (Fig. 5e).

387 *3.6. Further evidence for heterogeneous sources of noble gases on Earth*

388 From isotope geochemistry considerations, several distinct cosmochemical sources contributed
389 noble gases to the growing Earth.

390 (i) Solar neon has been identified in the mantle (Honda et al., 1991). It does not necessarily imply a
391 solar origin for heavier noble gases because the solar component is rich in light noble gases
392 compared to Chondritic, so that mixing between Solar and Chondritic may impact light noble
393 gases only. Neon is unlikely to have been trapped in cometary ice given its elevated condensation
394 temperature.

395 (ii) The isotopic compositions of H, N and mantle Kr isotopes are consistent with a chondritic
396 source.

397 (iii) Atmospheric xenon is unique in the solar system and cannot be derived directly from Solar or
398 Chondritic. The light Xe isotopes are isotopically fractionated (depleted) by 3-4 %/amu relative to

cosmochemical Xe (either Solar or Chondritic). However this mass fractionation may be secondary and may not be a source signature (Pujol et al., 2011; Avice and Marty, 2014). Even when corrected for mass-dependent fractionation, the atmospheric Xe isotope composition cannot be derived from either Solar or Chondritic, because it is depleted in the heavy isotopes ^{134}Xe and ^{136}Xe relative to the latter. Secondary (nuclear) processes such as fission of heavy radionuclides can only enrich, and not deplete, these isotopes. This observation led Takaoka (1972) and Pepin (1991) to postulate the existence of a distinct primordial Xe having sourced the Terrestrial atmosphere (the Xe-U component of Pepin, 1991). This mass-independent isotope fractionation of Xe isotopes might have resulted from an unknown nuclear effect, or from a different mix of nucleosynthetic components that were not fully homogenized during solar system formation.

Given the possible cometary origin of atmospheric noble gases, we speculate that this exotic Xe component could have been carried out from the outer solar system to its inner region by icy bodies. The dichotomy between mantle chondritic Xe and surface Xe would then be a record of Terrestrial growth, with the Proto-Earth being built from chondritic-like material, and its surface having been subsequently contributed by cometary bodies. In order to supply all atmospheric ^{36}Ar (5.55×10^{15} moles, Ozima and Podosek, 2002), a cometary contribution between 3×10^{21} and 6.5×10^{23} g would have been necessary (computed with a ^{36}Ar cometary concentration of $2.1 \pm 1.9 \times 10^{-7}$ mol/g, and a cometary density of 0.5 from the 67P measurements). This is in great agreement with the most recent estimates of the cometary flux to the Earth during the TLHB. Thus, there is no "missing comet bombardment" unlike the earlier conclusions of Marty and Meibom (2007) based on an higher ^{36}Ar content of comets. The impact of a single cometary body with a radius of 100-300 km, or of a limited number of smaller bodies, could have sufficient to supply all atmospheric argon. Therefore, it may be possible that the supply of atmospheric volatiles to the Terrestrial planets was a stochastic process involving a small population of objects, which might have resulted in an heterogeneous distribution of outer solar system isotopic signatures. In this respect, the high abundance of ^{36}Ar in the

atmosphere of Venus could effectively be the result of random contribution of cometary material ton inner planets.

3.7. Origin of prebiotic material on Earth constrained from atmospheric argon

Peptide synthesis from amino-acids appears a necessary step for the development of prebiotic material on Earth. Although a number of possible scenarios invoking catalytic reactions in specific Terrestrial environments have been advanced, it has also been proposed that impacts could have favored peptide synthesis from ET amino-acids (Blank et al., 2001). Sugahara and Mimura (2014) simulated comet impacts by shock experiments at low temperature in the range 5-26 GPa and found that linear peptide synthesis up to tripeptide forms could be achieved only with initially low temperature. As formation of linear peptide appears a key process for further elongation of peptide chains, these authors argued that comet impacts might have played a key role in the formation of prebiotic material on Earth. An upper limit for the total delivery of potential prebiotic material to Earth by comets can be set by considering that all atmospheric ^{36}Ar was delivered by icy bodies to the Earth's surface. Carbonaceous chondrites contain about 80 different amino-acids (Ehrenfreund and Charnley, 2000), with their total concentration amounting for 250 ppm (e.g., Martins et al., 2007). Their concentration in cometary matter is unknown. Studies of Stardust cometary grains suggest similarities with stratospheric IDPs and carbonaceous chondrites (Brownlee et al., 2006). Thus, for the purpose of illustration, we consider that the carbonaceous chondrite concentration is representative of cometary matter. The mass of cometary material having the ^{36}Ar content of 67P ($2.1\pm1.9 \times 10^{-7}$ mol/g) that would have delivered atmospheric ^{36}Ar (5.55×10^{15} moles) is $0.2\text{-}5 \times 10^{22}$ g, leading to the delivery of $0.6\text{-}13 \times 10^{18}$ g of amino acid-rich material. This amount is comparable to the present-day mass of the biosphere ($\sim 2 \times 10^{18}$ g). With the present-day mass of the oceans, comets could have delivered up to 10-40 ppm prebiotic material to seawater. Thus depending on the efficiency of survival and of peptide synthesis upon impact, comets might have contributed non-negligible prebiotic material during the TLHB.

449 *3.8. The case of the Martian atmosphere*

450 The composition of the atmosphere of Mars (Table 4) is reasonably known from the in-situ
451 analyses by the Viking and the Mars Science Laboratory (MSL) landers, and from the analysis of
452 some of the Martian meteorites which contain atmospheric volatiles trapped in impact glasses (see
453 Bogard et al., 2001, for a comprehensive review, and Wong et al., 2013, and refs. therein, for recent
454 Curiosity MSL data). Compared to Earth and bulk meteorites, the Martian atmosphere is rich in ^{15}N
455 ($\delta^{15}N$= 572±82 ‰ relative to the Terrestrial atmosphere, Wong et al., 2013) and in deuterium (δD =
456 5,500 ‰ relative to the oceans; Bogard et al. , 2001; Montmessin et al., 2005, and refs therein; Fig.
457 2). Likewise, neon, argon (^{36}Ar, ^{38}Ar) and xenon isotopic signatures are enriched in their heavy
458 isotopes by 246, 151, and 33 ‰, respectively, relative to the solar compositions (Table 4). In
459 contrast, volatiles in the Mars interior, also analyzed in Martian meteorites, have isotopic ratios
460 comparable to inner solar system signatures (Fig. 2).

461 The atmospheric signatures have been interpreted as resulting from isotope fractionation during
462 atmospheric escape processes. Possibilities include solar wind-induced photochemical reactions
463 producing electrons able to dissociate N_2 for nitrogen (Mc Elroy et al., 1976), and sputtering for the
464 noble gases (Jakosky et al., 1994). Chassefière and Leblanc (2004) reviewed potential escape
465 mechanisms that might have affected the atmosphere of Mars and concluded that (i) not a single
466 process can account for the observations; and (ii) non-thermal processes are required and therefore
467 might have affected different atoms/species at variable levels independently of their masses.
468 Atmospheric loss is attested by morphological evidence on Mars indicating past water flows (Carr
469 and Head, 2003), and major escape processes might have taken towards the end of the Noachian
470 period 4.0-3.7 Ga ago. Thus escape-related fractionation of Martian volatiles is likely to have played
471 a key role in fractionating elemental abundances and isotopic compositions. Nevertheless, we
472 suggest here that possible source effects, namely cometary contributions, should be also considered,
473 especially after the major escape events that resulted in a tenuous atmosphere on Mars.

474 Several key observations are not entirely consistent with atmospheric escape as a cause of isotopic
475 enrichments. (i) The noble gas elemental abundance of the Martian atmosphere and the C/N ratio
476 (Pepin, 1991) are comparable to those of Venus and Earth (for which similar escape fractionations
477 are unlikely). (ii) The krypton isotopic composition of the Martian atmosphere is close to the solar
478 composition (Bogard et al., 2001, and refs. therein), contrary to those of Ne, Ar and Xe (Table 4).
479 Because the mass of Kr is intermediate betwen those of Ne, Ar and Xe, this difference points to other
480 processes than escape-related fractionation. We note that the Martian Ne isotopic composition is
481 within the range of chondritic values, e.g., the so-called Ne-A (planetary) component of
482 carbonaceous chondrites. Neon is not expected to be trapped in comets, so that Martian atmospheric
483 Ne could have been sourced by an another cosmochemical reservoir. The high $^{38}Ar/^{36}Ar$ ratio (0.238
484 ± 0.005; Atreya et al., 2013) is indeed a good diagnostic of escape-related fractionation, but cosmic-
485 ray production of noble gas isotopes (which also increases this ratio) could have been sigificant for
486 the tenuous Martian atmosphere. Such effects need to be assessed. The Xe isotope fractionation
487 could have arisen from a Xe-specific non-thermal escape process, like the one proposed for
488 Terrestrial atmospheric Xe (Pujol et al., 2011). (iv) The isotopes of C, O, S in the Martian surface
489 and crust are little fractionated. CO_2 is enriched in ^{13}C relative to Terrestrial ($\delta^{13}C$ = 46 ± 4 ‰;
490 Mahaffy et al., 2013) but to a much lesser extent than ^{15}N in nitrogen ($\delta^{15}N$= 572±82 ‰, Wong et al.,
491 2013). However this argument is not carved in stone since C, O, S isotope compositions might have
492 been buffered by crustal reservoirs.

493 We suggest that some of the Martian atmospheric characteristics are consistent with a cometary
494 origin. This possibility was proposed by Owen and Bar-Nun (1995) who noted $^{36}Ar/^{84}Kr/^{132}Xe$
495 similarities between the Earth and Mars despite large differences in atmospheric pressures,
496 suggesting that a non-fractionating process supplied atmospheric noble gases to both planets. They
497 argued that the Martian $^{36}Ar/^{84}Kr/^{132}Xe$ ratios resemble to those obtained in experimentally grown
498 ice, and proposed impacts of icy planetesimals as the source of Martian surface volatiles. Although
499 we do not think that the Xe abundance pattern and its isotopic composition are cometary (the Martian

500 Xe fractionation could result from Xe-specific non-thermal processing like on Earth, Pujol et al.,
501 2011), we test below a possible cometary contribution by using other volatile element signatures.

502 First, in a N vs. H isotope diagram, the atmosphere of Mars plots close to the field of comets (Fig.
503 2). Second, the Martian C/N, C/^{36}Ar and N/^{36}Ar ratios are all close within a factor of 2 to our estimate
504 for cometary composition (and to Earth's surface inventory, Table 4, Fig. 6), whereas the C/^{36}Ar and
505 N/^{36}Ar ratios are markedly different from the chondritic values (Table 4). The δ^{15}N vs. $^{36}Ar/N_2$
506 isotope variations are consistent with a two-component mixing (Fig. 7). In this format, mixings
507 define straight lines (curves in the semi-log diagram of Fig. 7). Both the Martian atmosphere's δ^{15}N
508 and $^{36}Ar/N_2$ values are within the estimated range of cometary ratios, and variations observed in
509 Martian meteorites define another end-member having lower δ^{15}N and higher $^{36}Ar/N_2$ ratios, that
510 could be chondritic and/or terrestrial contamination (preferential adsorption of nitrogen relative to
511 argon). The Kr isotope composition is solar-like, as expected for a cometary origin. After correction
512 for mass-dependent fractionation, Xe is also isotopically solar (Pepin, 1991).

513 We then explore the possibility of a cometary contribution by scaling potential cometary fluxes on
514 Earth to the dimensions of Mars. In Fig. 6, the impacting fluxes on Earth have been adapted to the
515 Martian dimensions by dividing them by a factor of ~5. This factor is obtained from the ratio of
516 planetary surfaces (3.5 for Earth/Mars) plus a gravitational focusing effect of ~1.5. It comes out that
517 a Martian late veener would supply far too much cometary volatiles to Mars (Fig. 6). For a TLHB
518 total (asteroidal+ cometary) flux of ~2 x 10^{23} g, the total Martian LHB (M-LHB) flux would be ~5 x
519 10^{22} g. Ten % comet in the M-LHB as on Earth gives a cometary flux on Mars of ~5 x 10^{21} g, which
520 is remarkably comparable to the range of 7-14 x 10^{21} g obtained from recent simulations. However, a
521 10% cometary M-LHB would add too much volatiles to Mars by ~1 order of magnitude (Fig. 6).
522 Thus either the LHB flux was heterogeneous between Earth and Mars, or part of the Martian
523 atmosphere was subsequently lost after M-LHB by atmospheric escape processes. In line with the
524 second possibility, the D/H values of the Martian interior and Martian atmosphere are shifted

525 towards D-enrichments (right-hand side in Fig. 2) relative to inner solar system values defined by
526 Earth and chondrites on one hand, and by comets on another hand. Thus the Martian D/H value is too
527 high to be solely cometary, and it is likely that atmospheric escape processes played a role in
528 fractionating hydrogen isotopes, and, possibly, other light species (neon ?). Finally, the continuous
529 flux of IPDs on Mars could constitute a non-neglible source of cometary volatiles (Fig. 6), if made
530 predominantly of cometary material. In line with this possibility, Fries et al (2016) proposed
531 recently that "meteor showers" made of cometary debris could account for the presence of methane
532 in the Martian atmosphere.

533 **4. Conclusions and outlook**

534 The in-situ analysis of coma from comet 67P together with remote sensing data from other comets
535 gives semi-quantitative constraints on the origin of volatile elements in Terrestrial planets.
536 Particularly important is the discovery of significant amounts of primordial argon in the coma of
537 67P, well above the level observed in primitive meteorites.

538 For Earth, the major volatiles documented here - H_2O, C, N - were sourced mostly by material
539 akin of wet asteroids. Only a few percents at most of water, carbon and nitrogen could have
540 originated from comet-like bodies. However the terrestrial atmosphere could contain a significant,
541 possibly dominant, fraction of cometary noble gases, which might have been delivered during the
542 Terrestrial Late Heavy Bombardment around 4 Ga ago. This possibility is in line with dynamical
543 models of solar system evolution, which predict a significant fraction of icy bodies in the TLHB.
544 Resolving this question will require a better estimate of the volatile element inventory of the Earth.
545 Models that ascribe a "low" volatile content of our planet (Halliday, 2013; Dauphas and Morbidelli,
546 2014) imply that atmospheric noble gases are cometary, whereas the "high" volatile content model
547 (Marty, 2012) does not require comets and is consistent with a predominantly asteroidal (chondrtitic)
548 source (see Fig. 4b). A chondritic, rather than cometary, origin for atmospheric noble gases is
549 independently consistent with Ne-Ar isotope variations among the major terrestrial reservoirs

including the atmosphere (Marty, 2012). This problem may appear too specialized for a non-specialist audience, but there are outstanding issues beyond it, such as the origin of prebiotic material on Earth (which could be traced with noble gases), the early processes of volatile delivery and escape, the timing of volatile delivery (which can be addressed with noble gas radiogenic isotopes), and the composition and mode of transfer of material from the outskirts of the Solar System. For example, the isotopic composition of atmospheric xenon cannot be simply related to either the solar nebula or asteroidal material. It requires a specific composition (U-Xe; Pepin, 1991) that could trace nucleosynthetic heterogeneities in outer solar system bodies. Resolving these important issues will require a better estimate of the volatile content of the mantle on the one hand, as well as precise cometary noble gas data including isotopes on the other hand.

A cometary contribution to Martian surface volatiles (nitrogen and noble gases) appears clearly possible, given our present state of knowledge, and is consistent with solar system dynamical models. However, the strong deuterium enrichment of the Martian surface can hardly be the result of a cometary contribution, and requires isotope fractionation during escape processes at least for these light species. Resolving source effects versus processing is a major challenge that will require, as for the Earth, a better knowledge of the Martian and cometary signatures for major and minor volatile species.

The volatile characteristics of the inner planets are consistent with trapping of protosolar nebula gas during early stages on accretion, and contributions from wet bodies from increasing heliocentric distance during the sequence of planetary building and late accretion. However no unified picture emerges for the origin(s) and processes of delivery of inner planets' atmospheric volatiles. What are now highly needed are measurements of the noble gas abundances in comets and the isotopic ratios of neon, krypton and most importantly xenon. Such measurements should be given a high priority in future cometary missions. The requested level of precision (better than percent for noble gas isotopic

574 ratios, and per mil for stable isotopes including the three isotopes of oxygen) will require a cometary
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781

	lower	upper	mean	+/-	
Water			100		*1*
Carbon					
CO_2	2.5	80	41	39	*2*
CO	1	30	16	15	*2*
CH_4	0.6	1.5	1.05	0.45	*2*
C_2H_2	0.1	0.46	0.28	0.18	*2*
C_2H_6	0.4	10.5	5.45	5.05	*2*
others	0.4	2.4	1.40	1.00	*2*
ΣC			71	65	
Nitrogen					
NH_3	0.3	1.5	0.90	0.60	*2*
HCN	0.1	0.6	0.35	0.25	*2*
N_2			0.088	0.083	*3*
ΣN			1.43	1.02	
$N_2/\Sigma N$			0.06	0.07	
Argon					
^{36}Ar			1.20E-03	1.10E-03	*4*

Table 1: Coma composition. *1*: Species and elements are molar fractions normalized to water (=100). *2*: The range of C- and N-bearing species concentrations are for all comets analyzed so far (compiled from Bockelée-Morvan, 2011; Mumma and Charnley, 2011, Le Roy et al., 2015). *3*: The N_2 content is derived from the comet 67P measurement by Rubin et al. (2015), the good correlation between CO and N_2, and the cometary CO content given above. *4*: Ar/H_2O ratio from Balsiger et al. (2015).

	Coma		Bulk	
	mole/g	+/-	mole/g	+/-
^{12}C	0.012	0.011	0.037	0.028
^{14}N	0.00025	0.00018	0.0018	0.0010
H_2O	0.017	0.0010	0.019	0.0010
^{36}Ar	2.07E-07	1.89E-07	2.07E-07	1.89E-07

Table 2: Cometary bulk composition. Abundances are computed with data from cometary composition (see text), Table 1 and refs. therein. The ^{36}Ar content is for 30% water ice.

	Atmosphere	Halliday, 2013 - HA				Marty, 2012 - MA				Dauphas & Morbidelli, 2015
		Mantle	atm.+mantle			Mantle	atm.+mantle			
	mol/g	mol/g	+/-	mol/g	+/-	mol/g	+/-	mol/g	+/-	DM, mol/g
^{12}C	1.29E-06	4.00E-06	1.89E-06	5.29E-06	1.89E-06	6.38E-05	2.50E-05	6.51E-05	2.50E-05	5.60E-06
^{14}N	5.98E-08	7.48E-09	4.71E-09	6.73E-08	8.97E-09	9.08E-08	4.60E-08	1.51E-07	4.60E-08	1.12E-07
H_2O	1.48E-05	1.80E-05	6.00E-06	3.28E-05	6.00E-06	2.00E-04	9.60E-05	2.15E-04	9.60E-05	4.35E-05
^{36}Ar	9.33E-13	3.03E-15	1.59E-15	9.36E-13	1.59E-15	7.83E-14	4.30E-14	1.01E-12	4.30E-14	9.29E-13

Table 3: Earth's volatile composition. Data are in mol/g for bulk Earth (mass = 5.98 x 10^{27} g). "Atmosphere" refers to the atmosphere *sensu stricto*, the oceans and crustal and sedimentary rocks (data are from Ozima and Podosek, 2002, Dauphas and Morbidelli, 2014, and refs. therein). Uncertainties on the C and N contents of the "atmosphere" are ~20 %. Halliday (2013; HA) considered the mantle sources of MORBs and of ocean basalt islands (note that the listed abundances differ from those given by Halliday, 2013, in which some of the numbers could not be reproduced). Marty (2012; MA) computed volatile element abundances from K-Ar systematics which predicts that, for a K content of 280 ± 60 ppm (1 σ; Arevalo et al 2009), half of radiogenic ^{40}Ar produced by the decay of ^{40}K ($T_{1/2}$ = 1.25 Ga) over 4.5 Ga is still trapped in the mantle. The other volatile contents are computed from $^{40}Ar/N/C/H_2O$ ratio estimates for OIBs and MORBs. Dauphas and Morbidelli (2014; DM) published recently estimates for volatile content of inner planets and other cosmochemical reservoirs, which are intermediate between HA and MA abundances, although closer to the former. HA and BM are assumed to represent two limiting cases for the Terrestrial inventory of volatile elements in the silicate Earth+surface.

	Mars atm. mol/g	Carb. chondr. mol/g	Comets mol/g
^{12}C	$8.6.10^{-10}$	$2.3.10^{-3}$	$3.7.10^{-2}$
^{14}N	$3.4.10^{-11}$	$6.4.10^{-5}$	$1.8.10^{-3}$
^{36}Ar	$9.1.10^{-15}$	$3.9.10^{-11}$	$2.1.10^{-7}$
C/N	25	36	21
$C/^{36}Ar$, 10^4	9.5	5,900	17.9
$N/^{36}Ar$, 10^3	3.7	1,640	8.7
$^{20}Ne/^{22}Ne$	7 - 10	8.5 - 12.5	?
$\delta^{15}N$ ‰	572 (±82)	-30 to + 150	+600 to +1,000
δD ‰	+5,500	-300 to +1,000	0 to 2,000
F_{Ne} ‰/u	246	173	?
F_{Ar} ‰/u	151	6	0
F_{Kr} ‰/u	±10	10	0
F_{Xe} ‰/u	33	3	0

Table 4: Volatile element composition of the Martian atmosphere (sensu lato, including estimates for surface reservoirs). Elemental abundance data are from Bogard et al. (2001), Dauphas and Morbidelli (2014), and refs. therein. Key elemental ratios and isotope data are also given, referenced in Bogard et al. (2001), Dauphas and Morbidelli (2014), Mahaffy et al. (2013), Atreya et al. (2013) and Wong et al. (2013). The $\delta^{15}N$ values are given in parts per mil (‰) relative to terrestrial atmospheric nitrogen. The δD values are global estimates from Montmessin et al. (2005), given as deviations in ‰ relative to terrestrial ocean water. The F_X values (where X is a noble gas isotope ratio), represent deviations (in ‰ per atomic mass unit) from the solar composition (Dauphas and Morbidelli (2014). Note that Ar and Xe appear highly fractionated relative to Solar and compared to Kr, the latter being close to Solar. Ar, Kr and Xe are assumed to be isotopically solar in comets, as cometary ice should have trapped quantitatively these elements from the protosolar nebula. Neon is unlikely to have been trapped in comets, given the inferred formation temperature of these objects, and its abundance and isotopic composition cannot be inferred.

843

844 **Figure Caption**

845 **Fig. 1**. (a) Comparison of the abundance patterns of noble gases in the Sun ("Solar", yellow
846 hexagons), in volatile-rich primitive meteorites ("CI", empty circles) and in the atmospheres of Earth
847 (black-filled circles) and Mars (red squares). Note that the latter are depleted in xenon relative to
848 krypton and meteorites by about a factor of 20, and that the Kr/Xe is close to the solar abundance
849 (the "missing" Xe for Earth's atmosphere is represented by the blue area, see text). Modified after
850 Porcelli and Ballentine (2002), Mars data from Dauphas and Morbidelli (2014). (b) Comparison of
851 the Xe isotope compositions of meteoritic Xe (empty circles, "Q-Xe"), atmospheric Xe
852 ("Atmosphere", black-filled circles), and Martian atmosphere Xe ("Mars-Xe", red squares),
853 expressed in deviations in parts per mil (‰) relative to the solar composition (derived from solar
854 wind measurements, yellow bar). In this format, the solar composition gives a flat spectrum at 0 ‰.
855 The meteoritic composition is slighlty different from Solar, whereas Martian and Terrestrial Xe are
856 clearly fractionated (enriched in heavy isotopes) compared to solar, prohibiting direct derivation
857 from the protosolar nebula and, by inference, from cometary ice (see text). Data from Ott (2014).

858 **Fig. 2**: Nitrogen vs. hydrogen isotope variations among solar system reservoirs (modified after
859 Aléon, 2010, and Füri and Marty, 2015). The H and N isotopic ratios are normalized to the
860 respective protosolar values. Fractionation trajectories are displayed for illustration. "Ion-molecule"
861 is from Aléon (2010), representing isotope fractionation during ion molecule exchange. "Kinetic"
862 illustrates the case of isotope fractionation proportional to the inverse square root of masses and is
863 only one of the possibilities (others involve reduced masses, molecular masses instead of atomic
864 masses etc.) Light blue dots: Giant planets; beige dots: inner planets and Titan; grey squares: bulk
865 primitive meteorites; light green diamonds: comets.

866 **Fig. 3:** $^{36}Ar/H_2O$ (*a*) and $^{36}Ar/N$ (*b*) versus D/H mixing diagrams between cometary and chondritic
867 (asteroidal) end-members. The mixing curves are constructed as $[X/Y]_{observed} = \alpha \times [X/Y]_1 + (1-\alpha) \times$
868 $[X/Y]_2$ for two end-members having cometary/chondritic elemental/isotopic compositions.

869 Coefficients α are computed in each case using the end-member compositions (Tables 1 and 2).
870 Terrrestrial reservoir compositions from Table 3: HA, DM and MA are the bulk Earth estimates from
871 Halliday (2013), Dauphas and Morbidelli (2014) and Marty (2012), respectively. Dotted lines
872 indicate the percentage of cometary water (a) and of cometary nitrogen (b), respectively, on Earth.

873 **Fig. 4**: D/H versus ^{36}Ar/C mixing diagram between cometary and chondritic (asteroidal) end-
874 members. HA, DM and MA are the bulk Earth estimates from Halliday (2013), Dauphas and
875 Morbidelli (2014) and Marty (2012), respectively. *a:* Dotted lines indicate the percentage of
876 Terrestrial carbon of cometary origin. *b:* Dotted lines indicate the percentage of cometary ^{36}Ar on
877 Earth.

878 **Fig. 5:** Abundances of C, N, H_2O and ^{36}Ar in the atmosphere *sensu largo* (air+oceans+sediments),
879 normalized to Chondritic (data from Tables 2 and 3). For comparison bulk Earth inventories from
880 Table 4 are also represented. *5a*: Effect of addition of a late veener (0.3 % of the Terrestrial mass)
881 made of 100 % comet. The ^{36}Ar inventory of the atmosphere (the main reservoir of noble gases) is
882 overabundant by several orders of magnitude, and major volatiles are also dominated by cometary
883 material. *5b*: same as *3a* but for a tiny contribution of cometary material in the late Veener,
884 arbitrarily chosen as 0.2 % to fit the Terrestrial abundances. *5c*: Effect of contribution of a Terrestrial
885 Late Heavy Bombardment (2×10^{23} g) made of 100 % comet. Given the uncertainties, the ^{36}Ar
886 content of the atmosphere can be matched, but the major volatiles at the Earth's surface are
887 dominated by a cometary component. *5d*: A~10% cometary contribution in the TLHB fits well the
888 ^{36}Ar inventory of the atmosphere, without impacting the isotope compositions of C, N and H. *5e*:
889 Effect of addition of a continuous IDP flux since 3.5 Ga ago. This flux has no impact on the
890 Terrestrial inventory of H, C, N and ^{36}Ar.

891 **Fig. 6**: Elemental composition of the Martian atmosphere, compared to Earth's atmosphere and
892 comets. Data sources: Pepin, 1991; Bogard et al., 2001; Dauphas and Morbidelli (2014). Water is not

893 represented here since its Martian inventory is unknown. The effects of a Martian late veneer (M-
894 LV), a Martian Late Heavy Bombardment (M-LHB) and continuous IDP flux are also represented.

895 **Fig. 7:** N isotope vs. $^{36}Ar/N$ ratio mixing diagram between carbonaceous chondrite (CCs) and comet
896 end-members. The dotted curves represent mixing trajectories between Comet and CCs on one hand,
897 and between MSL measured Martian atmosphric composition on another hand. between Viking and
898 Martian meteorite data are from Bogard et al. (2001) and refs. therein, and Chennaoui-Aoudjehane et
899 al. (2012). MSL data are from Wong et al. (2013) and Mahaffy et al. (2013).

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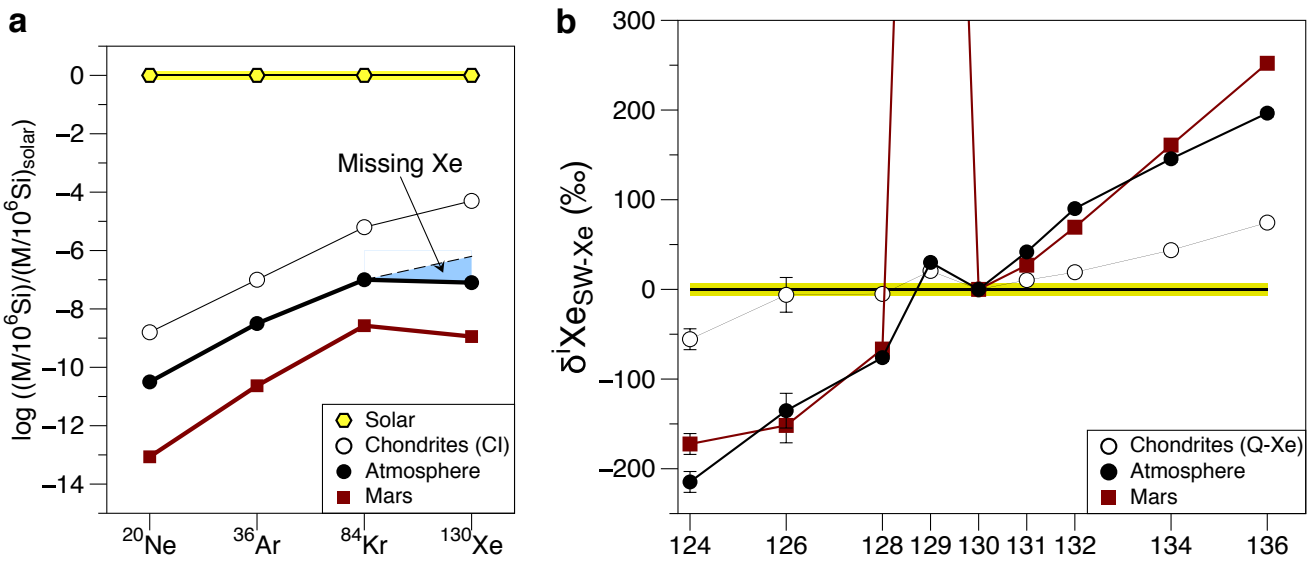
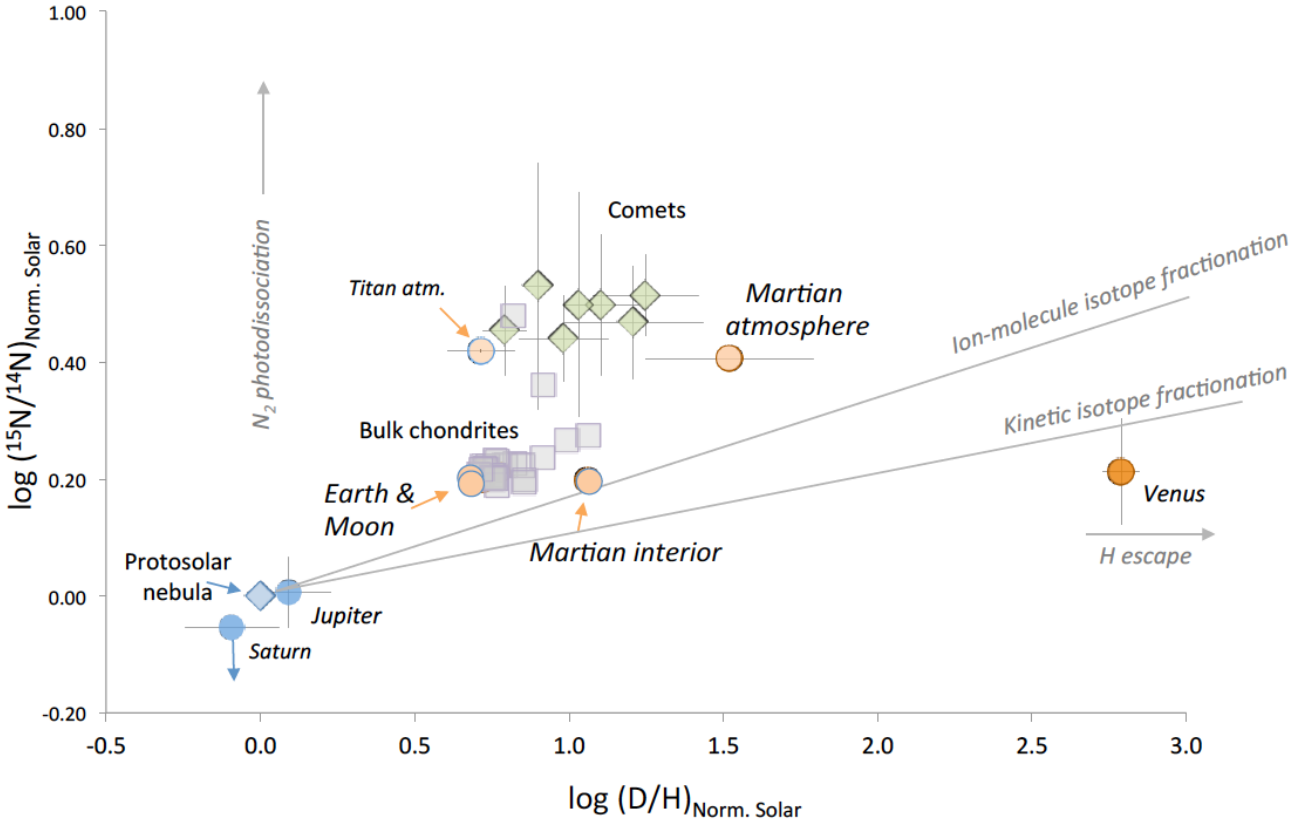

a
log ((M/10^6Si)/(M/10^6Si)solar)
0
−2
−4
−6
−8
−10
−12
−14
Missing Xe
Solar
Chondrites (CI)
Atmosphere
Mars
20Ne
36Ar
84Kr
130Xe
b
δ^iXeSW-Xe (‰)
300
200
100
0
−100
−200
Chondrites (Q-Xe)
Atmosphere
Mars
124
126
128
129
130
131
132
134
136

Fig. 1



910

911 **Fig. 2**

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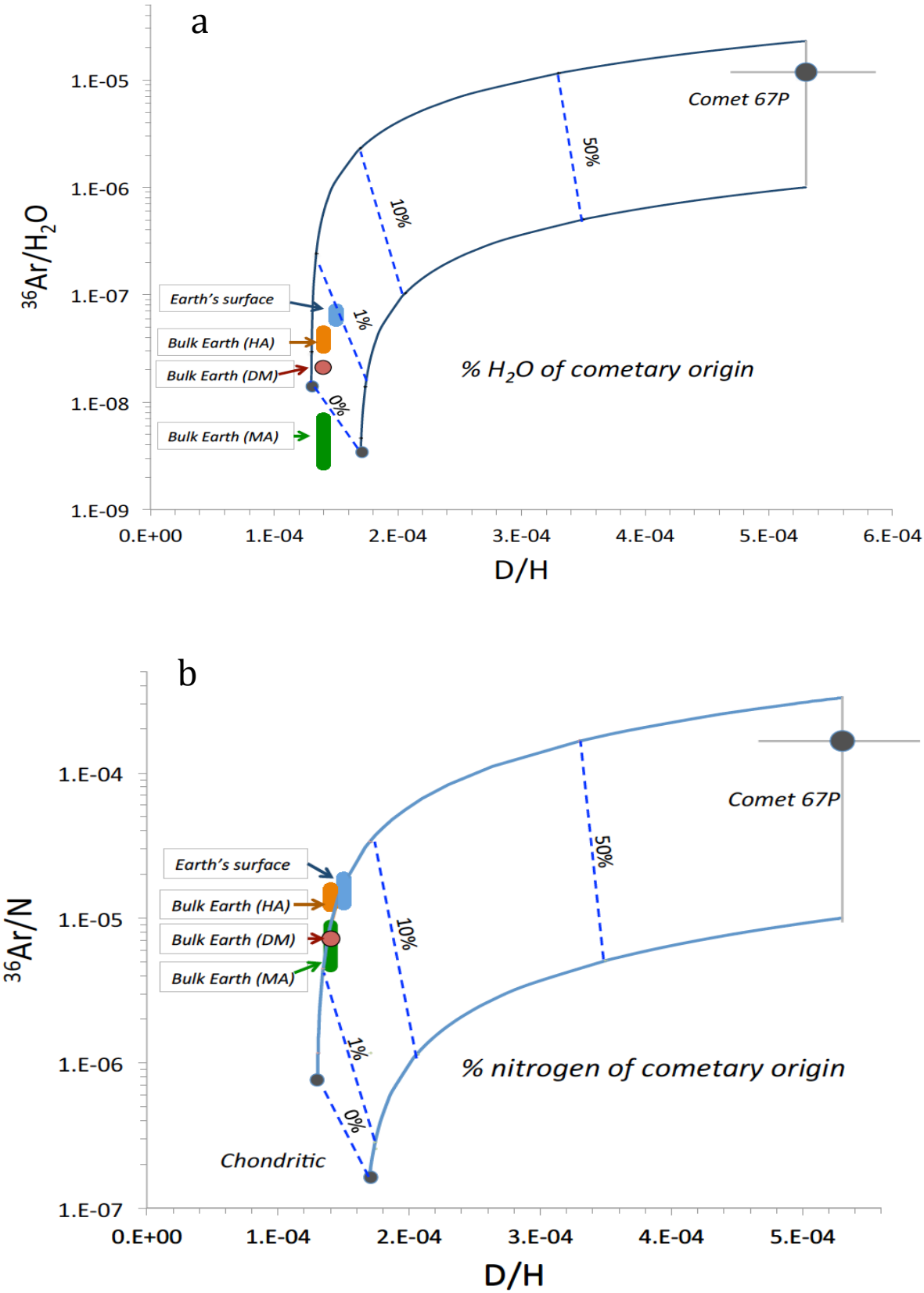

a
1.E-05
1.E-06
1.E-07
1.E-08
1.E-09
$^{36}Ar/H_2O$
0.E+00
1.E-04
2.E-04
3.E-04
4.E-04
5.E-04
6.E-04
D/H
Comet 67P
50%
10%
1%
0%
Earth's surface
Bulk Earth (HA)
Bulk Earth (DM)
Bulk Earth (MA)
% H_2O of cometary origin
b
1.E-04
1.E-05
1.E-06
1.E-07
$^{36}Ar/N$
0.E+00
1.E-04
2.E-04
3.E-04
4.E-04
5.E-04
D/H
Comet 67P
50%
10%
1%
0%
Earth's surface
Bulk Earth (HA)
Bulk Earth (DM)
Bulk Earth (MA)
% nitrogen of cometary origin
Chondritic

916

917 **Fig. 3**

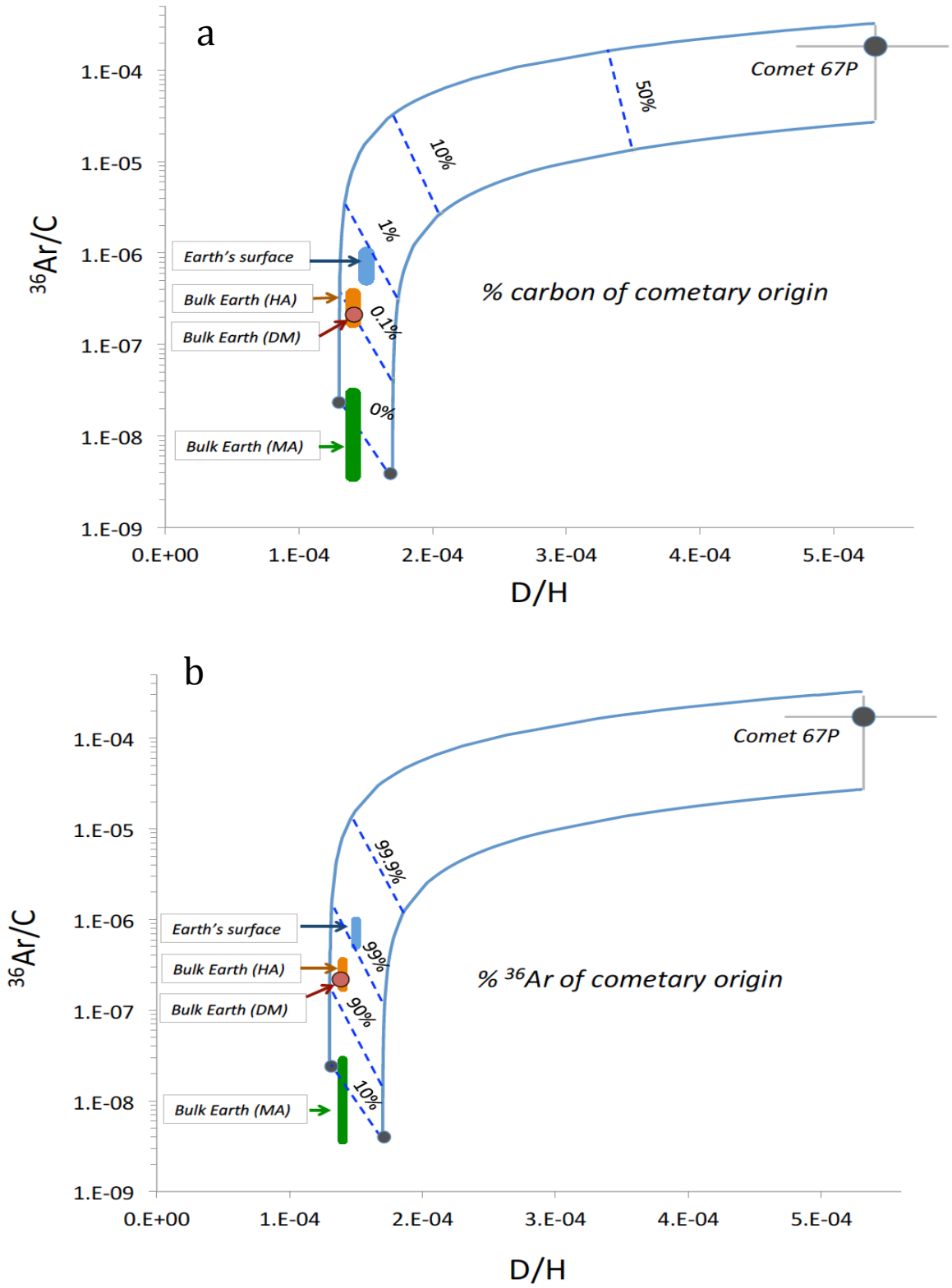

a
Comet 67P
50%
10%
1%
0.1%
0%
Earth's surface
Bulk Earth (HA)
Bulk Earth (DM)
Bulk Earth (MA)
% carbon of cometary origin
36Ar/C
D/H
1.E-04
1.E-05
1.E-06
1.E-07
1.E-08
1.E-09
0.E+00
1.E-04
2.E-04
3.E-04
4.E-04
5.E-04
b
Comet 67P
99.9%
99%
90%
10%
Earth's surface
Bulk Earth (HA)
Bulk Earth (DM)
Bulk Earth (MA)
% 36Ar of cometary origin
36Ar/C
D/H

Fig. 4

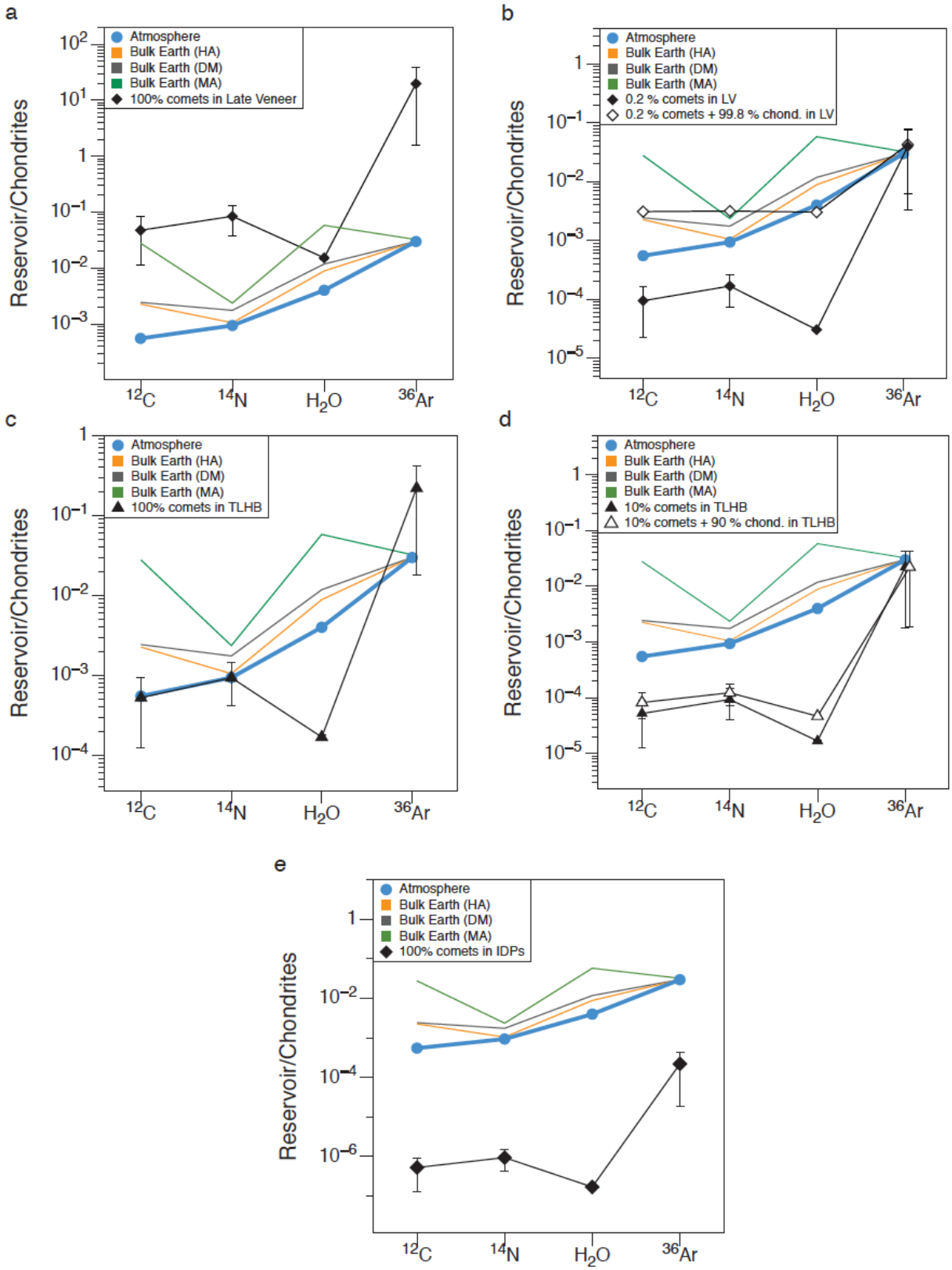
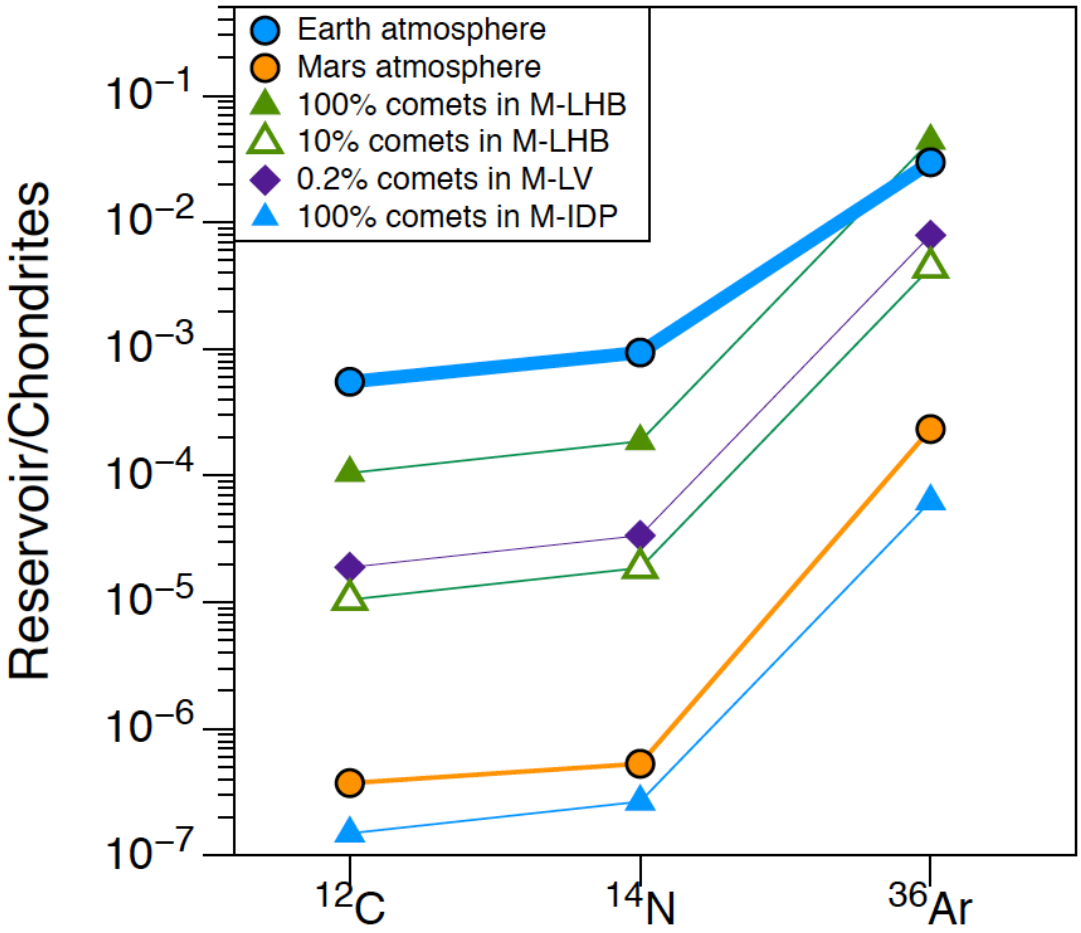

a
Atmosphere
Bulk Earth (HA)
Bulk Earth (DM)
Bulk Earth (MA)
100% comets in Late Veneer
Reservoir/Chondrites
10^{2}
10^{1}
1
10^{-1}
10^{-2}
10^{-3}
^{12}C
^{14}N
H_2O
^{36}Ar
b
Atmosphere
Bulk Earth (HA)
Bulk Earth (DM)
Bulk Earth (MA)
0.2 % comets in LV
0.2 % comets + 99.8 % chond. in LV
Reservoir/Chondrites
1
10^{-1}
10^{-2}
10^{-3}
10^{-4}
10^{-5}
^{12}C
^{14}N
H_2O
^{36}Ar
c
Atmosphere
Bulk Earth (HA)
Bulk Earth (DM)
Bulk Earth (MA)
100% comets in TLHB
Reservoir/Chondrites
1
10^{-1}
10^{-2}
10^{-3}
10^{-4}
^{12}C
^{14}N
H_2O
^{36}Ar
d
Atmosphere
Bulk Earth (HA)
Bulk Earth (DM)
Bulk Earth (MA)
10% comets in TLHB
10% comets + 90 % chond. in TLHB
Reservoir/Chondrites
1
10^{-1}
10^{-2}
10^{-3}
10^{-4}
10^{-5}
^{12}C
^{14}N
H_2O
^{36}Ar
e
Atmosphere
Bulk Earth (HA)
Bulk Earth (DM)
Bulk Earth (MA)
100% comets in IDPs
Reservoir/Chondrites
1
10^{-2}
10^{-4}
10^{-6}
^{12}C
^{14}N
H_2O
^{36}Ar

Fig. 5



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929

930 **Fig. 6**

931

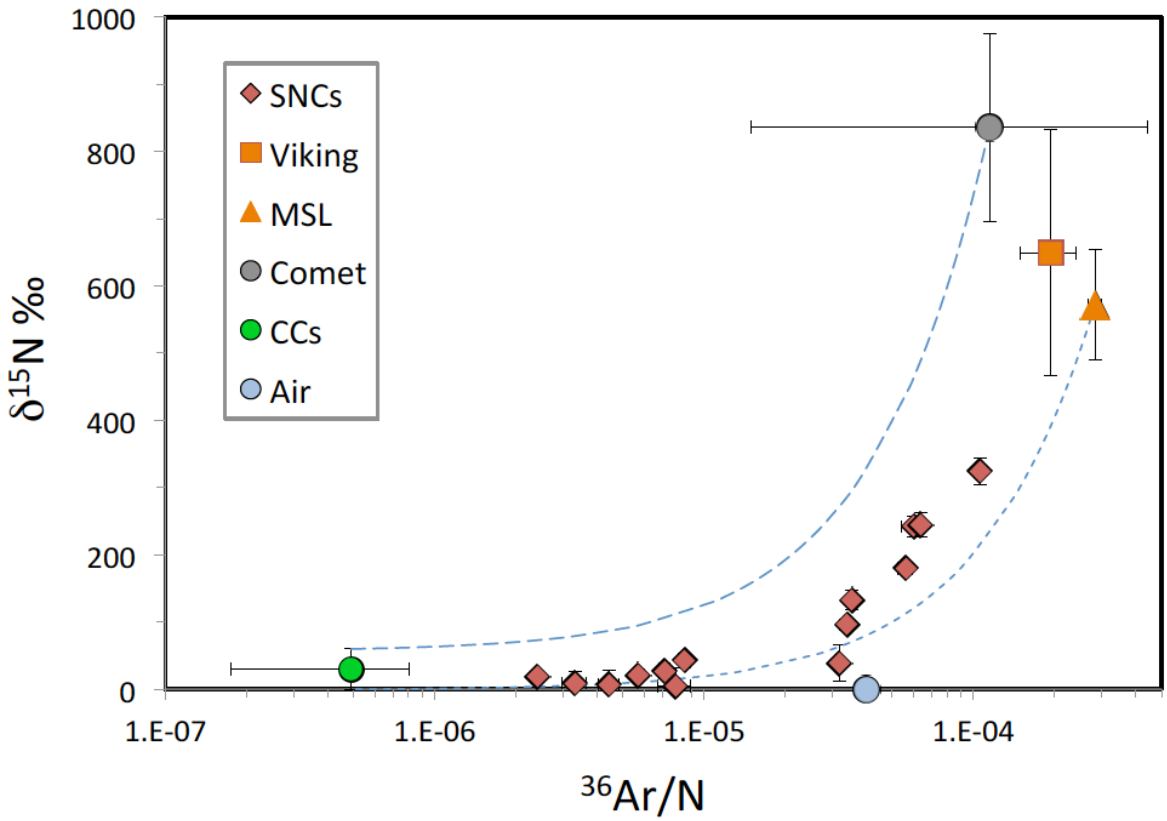
1000
800
600
400
200
0
δ15N ‰
SNCs
Viking
MSL
Comet
CCs
Air
1.E-07
1.E-06
1.E-05
1.E-04
36Ar/N

Fig. 7